On Antarctic Intermediate Water Mass Formation in Ocean General Circulation Models

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Abstract

Antarctic Intermediate Water is formed at the high mid-latitudes of the Southern Ocean. In many ocean general circulation model simulations with coarse resolution and a z-coordinate, a mid-depth salinity minimum characteristic of this intermediate water is reproduced. However, for the real ocean it remains unclear which are the dominant processes in the formation of this water mass and which are the source regions for contributing surface waters. To elucidate such processes and quantify intermediate water formation rates, two experiments with an ocean general circulation model were conducted. In one experiment, the traditional parameterization of horizontal and vertical mixing was applied, while the second model included the Gent-McWilliams parameterization for an eddy-induced transport velocity. In the latter application, the production and meridional export of intermediate water was found to be larger than in the first experiment. Furthermore, mid-latitude convective mixing, which had been argued to be the main intermediate water mass formation mechanism was found to be not as important as in previous model results. Passive tracer experiments indicated that diapycnal mixing and circumpolar subduction might also play important roles as water mass formation processes in this particular ocean circulation model.

1. Introduction

The distribution of Antarctic Intermediate Water (AAIW) is circumpolar in the Southern Ocean and can be identified from observations by an associated salinity minimum in a depth range of about 800-1000 m (McCartney 1977; 1982). McCartney (1977) argued that AAIW is formed by mid-latitude convection in the extreme eastern South Pacific Ocean and that Subantarctic Mode Water (SAMW) from that region is its precursor. In contrast, Sverdrup et al. (1942) originally proposed that the circumpolar extent of AAIW is indicative of subduction or along isopycnal transport that operates along the Polar Front Zone (PFZ). Convection and subduction are two of only three physical processes that may lead to the formation of a water mass like AAIW (Tomczak 1999). The third process is subsurface mixing which is less dependent on surface heat and freshwater fluxes.

Several ocean general circulation models (OGCM) have been used in recent years to study the formation of AAIW in the Southern Ocean. England et al. (1993) found that a region of oceanic heat loss correlated with a band of mid-latitude convection in the South Pacific Ocean. Maximum depths of convection were located in the extreme eastern South Pacific Ocean, the origin of a Southern-Ocean-wide, salinity minimum at about 800-1000 m depth in both observed (e.g. Levitus 1982) and modeled data. Mid-latitude convection was identified as the principal water mass formation mechanism in that particular coarse-resolution model. Ribbe and Tomczak (1997) carried out a similar diagnosis for a high resolution OGCM. They argued that property transport within the

Ekman surface layer across the main fronts, which confine the flow of the Antarctic Circumpolar Current (ACC), may contribute to AAIW characteristics and may even drive mid-latitude convection in OGCMs. In a subsequent study, Ribbe (1999) demonstrated in a series of computational experiments using an OGCM that changes in surface wind stress and northward Ekman transport in the surface layer lead to changes in the depth of mid-latitude convection.

The simulation of an AAIW salinity minimum is also sensitive to the parameterization of surface heat and freshwater fluxes; bulk flux boundary conditions yield more realistic results than traditional restoring boundary conditions (Large et al. 1997; Gent et al. 1998). Large et al. (1997) forced an OGCM with traditional restoring and alternatively with flux boundary conditions and used either a traditional parameterization for vertical mixing or the KPP vertical mixing scheme (Large et al. 1994). In the latter case, vertical mixing is non-uniform and parameterizes the effects of three active interior ocean mixing processes, i.e. breaking of internal waves, vertical shear instability, and convection. Through different combinations of surface forcing and vertical mixing parameterizations, Large et al. (1997) found that the simulation of the AAIW minimum is more sensitive to the formulation and parameterization of air-sea fluxes than to the interior vertical mixing formulation.

The implementation of an eddy parameterization scheme (Gent and McWilliams 1990; to be referred to as GM below) in z-coordinate OGCMs resulted in improved representations of temperature and salinity distributions in the thermocline and in a reduction of excessive deep convection at the higher latitudes of the Southern Ocean. The GM parameterization represents a "bolus transport" by mesoscale eddies. These are common features of the oceanic circulation in many regions of the Southern Ocean. Evidence of the role of eddy-subduction of AAIW has come from several recent studies. Marshall (1997) demonstrated eddy-associated subduction under idealized non-zero net surface buoyancy forcing. Lee et al (1997) used an eddy-resolving, isopycnal channel model to demonstrate net meridional transfer of tracers across a zonal jet due to the bolus transport by eddies which are associated with baroclinic instability of the jet (analogous to cross-ACC mixing). Wunsch (1999) discussed the size of eddy heat fluxes in the Southern Ocean as estimated from satellite altimetry, current meters and air-sea transfer studies. The zonal integration of these fluxes yields a large value for a net poleward eddy flux.

The GM parameterization of eddy-induced transports resulted also in a more realistic penetration of geochemical ocean tracers in OGCMs (England 1995; Duffy et al. 1997). In these applications, it is also evident that convective activity is almost completely absent in the mid-latitude regions. The reduction in high and mid-latitude convection was initially found by Danabasoglu et al. (1994) who simulated a much improved, global temperature distribution using a model with GM parameterization and isopycnal mixing. This is similar to findings by Hirst and Cai (1994) who demonstrated that when isopycnal mixing is used in OGCMs, convection is largely replaced by along-isopycnal mixing at or near the maximum allowable slope for density surfaces. This indicates that AAIW may be produced by direct subduction rather than convection in these new generation, z-coordinate OGCMs, which is consistent with the Sverdrup et al. (1942) original hypothesis based upon observations. In another recent study of AAIW formation in an isopycnal coordinate OGCM, direct subduction away from the surface was found to be the main AAIW formation process (Marsh et al. 2000).

The correct representation of AAIW in coarse resolution OGCMs is of great importance. This water mass probably plays a key role in the global thermohaline circulation, as North Atlantic Deep Water (NADW) must mix to intermediate density (i.e. become AAIW) before returning to the Atlantic Ocean, either directly at mid-depth (Rintoul 1991) or, after further mixing, as light surface water (Gordon 1986). While the distribution of AAIW has been discussed in new generation (GM-implemented), z-coordinate OGCM applications (e.g. Large et al. 1998; Gent et al. 1998), a diagnosis of the actual formation processes of this important water mass has been missing. Here, we address this issue and elucidate and contrast formation processes of AAIW in an OGCM with and without the GM parameterization scheme. For this purpose, an ideal ocean tracer was added to a coarse resolution OGCM. This allowed the passage of AAIW to be followed throughout the ocean model and the processes responsible for the formation of this water mass to be identified. In the model run with the GM parameterization – the run that yielded the most realistic water mass properties - mid-latitude subduction is identified as the main removal mechanism for Southern Ocean surface water. Furthermore, it is found that southern Antarctic surface water is a contributor to the intermediate water mass characteristics found to the north of the ACC.

2. Method

The model is based upon Modular Ocean Model 2 (MOM 2) with a setup similar to that of England et al. (1993). It has coarse horizontal resolution with 3.75 x 4.5 degrees, but quite high vertical resolution with 32 layers ranging in depth from 50-265 meters. Such vertical resolution minimizes the effect of numerical instabilities (Weaver and Sarachik 1990) and was also chosen to better resolve isopycnal flows, which are critical for representing the formation of AAIW. With split-time stepping, velocity and tracers were updated using timesteps of 1 day and 0.5 hours respectively. A realistic bottom topography was adopted.

Two series of experiments were conducted in this study. One series used the traditional horizontal eddy diffusion parameterization of sub-grid-scale motions while the GM scheme was applied in the other (Gent and McWilliams 1990). These two series will be referred to as the CART (synonymous for horizontal and vertical mixing within Cartesian coordinates) and GM runs respectively. Depth-dependent, horizontal and vertical diffusion coefficients were applied in the CART run. They were chosen a priori from Bryan and Lewis (1979) who took them to reflect some general bulk features of mixing in the ocean. The vertical diffusivity ranges from 0.3×10^{-4} to 1.3×10^{-4} m^2s^{-1} from the surface to the bottom. This reflects the effects of decreased stratification and increased small scale turbulence near the bottom. The horizontal diffusivity ranges from 0.5×10^3 to 1.0×0^3 m^2s^{-1}, being largest near the more energetic surface layers. In the GM run, the same vertical diffusivity profile was applied whereby it acted as background diffusion for an isopycnal mixing tensor with a constant isopycnal eddy diffusivity of 1.0 x 10^3 m^2s^{-1}. The background vertical diffusion term describes diapycnal mixing in this run. The value for horizontal diffusivity under GM was zero. The thickness diffusivity of the GM parameterization had the constant value of 1.0 x 10^3 m^2s^{-1}. Constant vertical and horizontal eddy viscosity coefficients of 2.0 x 10^3 m^2s^{-1} and 2.5 x 10^5 m^2s^{-1}, respectively, were applied in both the CART and GM runs.

Restoring boundary conditions were used for temperature and salinity with relaxation times of 30 and 50 days, respectively. The surface layer in the model is relaxed to the monthly climatological values of Levitus (1982). An attempt was made to include the effect of sea ice in the model. During June to August of each year, the surface salinity at 75^o S was restored to a maximum value of 36.0 psu with a linear interpolation defining the surface salinity south of 65^o S. This salinity relaxation resulted in simulated surface salinity around 35.0 psu which is not too far away from the maximum observed salinity. This is assumed to provide a first order simulation of brine rejection.

The monthly climatological wind stress field of Hellerman and Rosenstein (1983) was applied as the surface boundary condition for the momentum equations. The upper ocean circulation is in general quite sensitive to wind stress forcing and the deeper ocean can also be affected (Rahmstorf and England 1997). The Hellerman and Rosenstein (1983) wind stress field is based on few observations, a problem greater still south of 30^o S which is the main area of interest in the present study. This wind stress field is, however, a standard surface boundary condition in MOM and has been applied in previous modeling studies (e.g. England et al. 1993). Using a deep ocean acceleration technique (Bryan 1984), the model was spun up from the Levitus (1982) climatology to a steady state prior to applying passive tracers. Steady state was assumed to be attained when the drift of the global mean temperature and salinity was smaller than 0.01 ^{o}C and 0.004, respectively, per 100 years of integration at each vertical level (England et al. 1993). Based upon this definition, the spin up was completed after 1200 model years.

The two processes believed to participate in the formation of AAIW have their surface origins at different locations. Subduction of AAIW occurs along the path of the Antarctic Circumpolar Current in high mid-latitudes. In contrast, convection leading to the formation of SAMW (the AAIW precursor according to McCartney, 1977) occurs at the southern edges of the subtropical gyres and to the north of the PFZ due to wintertime surface water cooling. This system lends itself to the use of different passive tracers (e.g. Cox 1989) in each of these zones to estimate the relative importance of each of these two processes in AAIW formation. Accordingly, the model domain was divided into three regions: i) a South Antarctic (SA) zone extending from Antarctica to several degrees north of the PFZ; ii) a North Antarctic (NA) zone from the northern SA boundary to 30^{o} S, and iii) a zone covering the remaining world ocean north of 30^{o} S. In this model, the boundary between the SA and the NA regions falls just to the south of the line of zero wind stress curl. On completion of spin up, separate passive tracers were continuously added at the surface in all zones (by re-setting fractional tracer concentration to 1.0 at each tracer time step). Time series of the tracer distribution permit conclusions to be drawn about the formation regions and processes for SAMW and AAIW.

3. Results

a) Equilibrium State

The simulated global circulation of both runs corresponds well with that simulated by other coarse resolution models using MOM (England et. al. 1993; England 1997; Duffy et al. 1997). There are well-defined Southern Hemisphere subtropical gyres in all basins. The average Indonesian Throughflow into the Indian Ocean has a strength of 16 and 17 Sv (1 Sv = 10^6 $m^3 s^{-1}$) for the GM and CART runs, respectively. These lie in the range of observational estimates of 10 - 20 Sv (e.g. Godfrey 1996). The ACC is well represented with mean transports through Drake Passage of 119 Sv and 138 Sv in the CART and GM runs, respectively. These values compare well with observational estimates of 110 - 150 Sv (Tomczak and Godfrey 1994). In the GM run, the eddy-induced velocity counteracts meridional mean flow in the ACC region which, in turn, reduces the strength of the Deacon cell centered at about 50^o S. The model with GM parameterization results in a meridional overturning that better corresponds with observed meridional transports in the Southern Ocean due to a suppression of this cell.

The globally-averaged profiles of simulated temperature, salinity, and density for both model runs are shown in Fig. 1 and compared to Levitus climatology. The GM run produces too cold bottom waters (see Fig. 1a), a common feature in models employing the GM parameterization. One reason for this may be the shallow formation and insufficient downward penetration of NADW (Duffy et al. 1997) in the Northern

Hemisphere. The GM parameterization nevertheless greatly improves the simulation of thermocline characteristics. Temperatures within the thermocline are lower and the stratification is increased compared to the CART run, in better agreement with upper ocean observations.

The globally-averaged salinity is shown in Fig. 1b. A common deficiency of OGCMs without enhanced wintertime salinity forcing near Antarctica, or similar ad hoc or more sophisticated sea-ice models, is their inability to produce AAIW (Duffy et al. 1997). In such models, waters near to and south of the Polar Front sink to the bottom rather than participating in the formation of AAIW. This is avoided in the present model since the enhanced salinity forcing at high latitudes increased deep stratification, thus allowing fresher surface waters to subduct as AAIW at the Polar Front (Duffy et al. 1997). Previous sensitivity studies have demonstrated that the GM parameterization improves the representation of water masses (e.g. Large et al. 1997). This is reflected in the present model results which show a well-defined minimum of low salinity in the GM run (Fig. 1b).

Consistent with previous modeling studies, the stratification simulated in the CART run is too weak and potential densities are too low (Fig. 1c). The simulated distribution is unrealistic, although the artificial salt flux and restoring to 36.0 psu employed for the southernmost grid point in this model increased the stratification and densities significantly compared to solutions without this feature (e.g. Duffy et al. 1997). The two simulations yield significantly different stratification in the Southern Ocean (not shown). The CART run exhibits much too steep isopycnal slopes compared to observations whereas the GM run provides a much more realistic representation of this dynamically-important feature. This improvement rests on the parameterization of the eddy-mixing of potential density layer thickness. In conclusion, the combination of GM parameterization and enhanced wintertime salinity forcing returned a realistic representation of AAIW, but it remains unclear which processes form this water in the model.

To quantify the formation processes and transport of SAMW and AAIW in both the CART and GM run, the South Pacific Ocean was chosen as a control volume which previous studies already identified as the source region for the densest variety of SAMW as a precursor for AAIW. The boundaries were chosen with 150^{o} E south of Australia in the west, 65^{o} W in the east (Drake Passage), and 30^{o} S in the north. In the following, volume budgets for total transports and individual transport processes such as advection, diffusion, and convection are analyzed.

The annually-averaged volume transports across the three sections were integrated in four potential density ranges (Table 1). The chosen potential density ranges are: (1) surface water (σ_t < 26.0); (2) upper thermocline water (26.0 < σ_t < 26.8); (3) intermediate water (26.8 <σ_t < 27.5); and (4) deep and bottom water (27.5 < σ_t). This allows a separation between transports associated with SAMW (upper thermocline water) and AAIW (intermediate water). The σ_t = 27.2 kgm^{-3} surface lies well within the core of

AAIW in both the CART and the GM run as well as in the Levitus (1982) data set. Thus, it is consistent to track AAIW using this potential density range.

The meridional net northward transports of about 18 Sv (CART run) and 19 Sv (GM run) in the South Pacific Ocean correspond approximately to those leaving through the Indonesian Passage. The difference is due to the outflow of about 1 Sv through Bering Strait, which is open in these experiments. Surface volume transports in the CART and GM runs are 2.8 and 4.0 Sv, respectively. Distinct differences are simulated for the subsurface density ranges. While the transport in the combined upper thermocline and intermediate water density range is about 10.4 Sv in both runs, only the GM simulation exhibits a net northward transport in intermediate water (AAIW) density range of 4 Sv. This result most likely reflects unrealistically large, high latitude convection in the Southern Ocean in the CART run. This effect is associated with excessive bottom water production in the Southern Hemisphere, a freshened and weakened stratification of the deep ocean, and in turn, an excessive penetration of mid-latitude convection in SAMW and AAIW formation regions.

While the GM run exports 4 Sv of intermediate water northward, zonal in- and outflow of this water in the GM run are equal with about 62 Sv. The net of 4 Sv export is a product of local water mass conversion, in particular the conversion of upper thermocline water into intermediate water in the South Pacific Ocean basin. Of the remaining 19 Sv of upper thermocline water entering the South Pacific in the west in this run, about 9 Sv exits in the east and about 10 Sv exits across 30^{o} S, 4 Sv as surface layer and 6 Sv as upper thermocline water. Generally, water in the CART run is less dense than in the GM run (see also Fig. 1), which impacts upon the transports within the individual density ranges. Zonal transports within the intermediate water range of the CART run, for example, are larger than in the GM run. In contrast, zonal volume transports in the deep and bottom water range are more than twice as large in the GM run than in the CART run.

The northward spreading of AAIW in a depth range of about 500-1000 m is evident from the meridional distribution of salinity shown for the Global Ocean in Fig. 2. The subsurface salinity minimum observed at about 30^o S (Fig. 2c) seems to originate at the surface south of 50^o S. The GM run clearly simulates this salinity minimum (Fig. 2b), while it is almost absent from the CART simulation (Fig. 2a)

As expected from previous modeling studies (Danabasoglu et al. 1994; England 1995; Duffy et al. 1997), the GM run has significantly less mid-latitude convection than the CART run (Fig. 3). The main reason for this is the increased stratification in the GM run. Mid-latitude convection in the CART run (Fig. 3a) deepens from the west to the east. This feature is found in both the Indian and Pacific oceans in the CART run. It has been correlated with heat fluxes and has been identified as the intermediate water mass formation process in previous modeling studies without the GM parameterization (e.g. England et al. 1993).

The GM run exhibits no deepening of the maximum convection depth across the basins but is characterized by a somewhat sporadic convection (or lack of it) at mid-latitudes. A weak maximum with convection down to at most 350 m depth is present in the central South Pacific. Buoyancy gain or loss due to heat and freshwater fluxes for the GM run are shown in Fig. 4. The total flux (Fig. 4a) is the sum of the buoyancy flux due to heat and salt fluxes (Figs. 4b, c). Due to the relaxation boundary condition on surface temperature and salinity, the surface buoyancy fluxes in the model contain contributions from both the local atmospheric forcing and the horizontal surface transports. It should also be noted that the surface boundary condition for salinity is only a 'virtual' salt flux: the model is forced to a salinity climatology rather than by prescribed freshwater fluxes. In the real world, no salt is exchanged between the ocean and atmosphere. Cooling of surface water advected southward in the South Pacific gyre circulation results in a large enough buoyancy loss for convection to occur in the western South Pacific in the GM run. As a consequence, the depth of convection gradually increases in the western South Pacific. Heat loss continues all the way across the basin resulting in surface buoyancy loss (Fig. 4b), but a counter-acting freshwater flux, associated in part with the northward-directed Ekman layer transport, has a stabilizing effect upon the surface water (Fig. 4c). The net result is a reduction in maximum convective depth in the eastern South Pacific Ocean. In contrast to the GM run, the underlying stratification in the CART run is weak enough to promote an eastward deepening of the convective layer across the Pacific Ocean. Thus, the depth of convection is set by a subtle interplay of surface buoyancy forcing and underlying stratification.

The simulated salinity minimum is a result of northward fresh surface water transport from the higher latitudes into the subsurface ocean. The vertical transport processes involved seem to be well represented in the GM simulation (Fig. 2), yielding a net northward export of intermediate water of about 4 Sv across 30^o S (Table 1). To highlight the differences between the CART and GM simulation in representing various transport processes, a budget for the individual transport processes (advection, diffusion, and convection) of salinity (synonymous for freshwater) into and out of the intermediate water mass density range was calculated for the equilibrium state of both runs. A similar approach was chosen, for example, by Duffy et al. (1995) to investigate the oceanic uptake of bomb radiocarbon in MOM. In this work, however, the processes participating in the formation of intermediate water are considered, and not the processes that act to change water mass properties in general. Nevertheless, the results found in the present study for the transport of salinity into and out of the intermediate water mass density range are similar to those reported by Duffy et al. (1995).

The salt budget for both CART and GM simulations is shown in Fig. 5. The sum of the first column (total advection which includes the bolus transport), the third column (total diffusive mixing which includes vertical diffusive mixing), and the fifth column (convective mixing) is zero, i.e. the total change of the salt contents in the intermediate water mass volume is constant, since the budget is computed for the equilibrium state of model. Total advective transports in the GM parameterization are reduced due to the inclusion of the eddy flux bolus transport parameterization (2). Vertical transport by diffusive mixing (4) is significantly larger in the GM than in the CART run, since the transport by isopycnal mixing is included in the former. As shown in previous studies (Danabasoglu et al. 1994; Duffy et al. 1995), the most striking feature of the GM parameterization is a significant reduction of the vertical transport by convective mixing (5). The removal (negative transport) of salt in the CART run corresponds to a freshening of the intermediate water mass density range [and also of the deep and bottom water density range not included in this budget, but see Fig. 1 and Fig. 2]. The inclusion of isopycnal mixing in GM reduces convective mixing (5) and increases the vertical transport by diffusive mixing (4). Convection mixing shown in Fig. 3b for the GM run reaches only to depths of about 300-350 m and does not extend into the intermediate water mass depth level.

Danabasoglu et al. (1994) found that the GM parameterization reduces temperatures at intermediate depths by increasing the overall vertical transport of high latitude cold surface water into the interior of the ocean northward. Since high latitude water is of low salinity, this parameterization also increases the transport of freshwater into the ocean interior. In the absence of deep reaching convection, a process that reduces deep ocean densities in the CART case, cold and fresh surface water is subducted into the ocean interior and the intermediate water mass density range. In the GM case, subduction is essentially an eddy-induced mass flux (i.e. the bolus transport). Subsequently, isopycnal mixing acts to transform the subducting waters by modifying its temperature and salinity characteristics, but not its density.

Fig. 6 shows the equilibrium distribution (end of 1200 years integration) of the fraction of SA and NA tracer in the upper 2000 meters of a section across the Pacific Ocean at 30^{o} S from both the CART (Fig. 6a) and GM (Fig. 6b) run. For the latter, almost all of the water in the upper thermocline (~ 200 – 700 m, σ_t ~ 26.0 - 26.8 kg m^{-3}) originated at the surface in the NA region (Fig. 6b, bottom). The NA tracer maximum at 300 – 500 m depths near 150^o W apparently marks SAMW formed in the model. This SAMW, however, is considerably warmer and saltier (12-14 ^{o}C, 34.9–35.1 psu) than the SAMW actually observed in the central South Pacific. McCartney (1977) reports values of about 7.8 ^{o}C and 34.5 psu. This reflects the relatively weak and shallow, mid-latitude convection (Fig. 2) available for the formation of SAMW in the GM run. Together with the equilibrium distribution of SA tracer across this section (6b, top), the results also show that AAIW, marked by the salinity minimum and the σ_t = 27.2 kg m^{-3} surface, consists of about 30-40 % SA and 60 - 70 % NA water in the steady state of the GM run. The significant admixture of NA water at this depth - despite weak and shallow convection in the NA zone – is due to vertical mixing in the model. In contrast, the contribution of SA water to intermediate water is much reduced in the CART run (Fig. 6a, top), and the NA tracer penetrated much deeper into the ocean than in the GM run (Fig. 6a. bottom), making the CART run warmer and saltier at intermediate depth levels.

b) Transient Tracer Distributions

Passive tracers are also a useful tool to study the transient entry of the tracer into the ocean, and thus, the actual pathway of surface water away from its surface origin. After both the CART and GM run reached equilibrium, the passive tracer distribution described above was reset to zero and the SA and NA tracers newly assimilated into the model for an additional 100 years. This period is a rough time scale for the AAIW ventilation process, about the time it takes for newly formed AAIW to reach the subtropics.

In Figure 7a, the temporal evolution of the zonally-averaged SA tracer distributions simulated in both the CART (left) and GM (right) run are shown. Snapshots in time are drawn for 5, 25, 50, and 100 years. The SA tracer clearly ventilates predominately into the deep and bottom ocean in both model experiments. Convective activity simulated for the high latitudes of the Southern Ocean in the CART run results in deep penetrative mixing of the tracer to about 1000-1500 m between the southern model boundary and about 50^o S. After 100 years, a weak signal of the tracer (i.e. the 5 % contour) indicates northward spreading in a depth range of about 1000-1500 m. There is no indication of a subductive or an isopycnal mixing process. This is contrasted by the temporal evolution of the SA tracer distribution during the GM run. At intermediate water depth, a passive tracer maximum is found in depth of about 500-1000 m and north of about 50^o S. It overlaps with the salinity minimum characteristic for AAIW (Fig. 2b). The passive tracer maximum clearly originates at the higher latitudes of the Southern Ocean tracing the origin of the low salinity minimum. This indicates, that southern, fresh, Antarctic surface water contributes to the water mass characteristics of simulated intermediate water, and that the GM parameterization in combination with isopycnal mixing is able to represent this process.

The northern Antarctic tracer (NA) predominantly ventilates into the upper thermocline in both simulations (Figure 7b). In the CART run, convective mixing (indicated through almost vertical contours of the tracer distribution) is the predominant vertical mixing process in the higher latitudes. At about 50^o-60^o S, for example, the tracer is displaced to a depth of about 1000 m and spreads from there horizontally and vertically, with upwelling into the upper ocean toward the equator indicated by upward sloping contours. The NA distribution in the GM run is distinctly different with a shallower penetration in the southern latitudes, and more rapid ventilation toward the Equator. Tracer contours tend to follow the slope of density contours (not shown) representing the subductive or along isopycnal transport process.

The temporal evolution of the horizontal passive tracer distribution during both the CART and GM run are shown in Fig. 8. The tracer was averaged for the intermediate water density class. Both tracers contribute to the water mass mixture in this density range (see also Fig. 6). The SA tracer (Fig. 8a) ventilates more rapidly northward in the GM run than in the CART run. There are also regional differences. For example, no SA tracer ventilates into the Indian Ocean in the CART run, while in the GM run the SA tracer ventilates the eastern portion of both the South Indian and Pacific Ocean and traces the circulation of the subtropical gyres in the Southern Hemisphere ocean basins. After 100 years, the CART run simulates a weak tracer maximum in the eastern South Pacific Ocean, indicating also a circulation into the South Pacific Ocean. Convective mixing reaches deepest in this region of the Southern Ocean (Fig. 3a). But at equilibrium, this weak ventilation process does not lead to an intermediate salinity minimum (Fig. 2) and a net meridional transport northward across 30^o S (Table 1).

The upper thermocline is also ventilated much more rapidly in the GM run than in the CART run (Fig. 8b). In both runs, there is a spreading of upper thermocline surface water away from the eastern portions of both the South Indian and South Pacific Ocean with the gyre scale circulation. In CART, the ventilation process is more localized than in GM since in the former case convection is the primary vertical transport process, which in both the Indian and Pacific Ocean is deepest in the east. As in the case of the SA tracer, the contribution of the NA surface tracer to upper thermocline water mass characteristics is larger in GM than in CART. It is interesting to notice that in both runs maximum tracer concentrations can be found in similar locations, although the main vertical transport processes are different, i.e. convective mixing versus along isopycnal transports (both the resolved mean flow and parameterized eddy fluxes).

4. Discussion and Summary

The maximum depth of convection in the GM run is about 350 m, which is considerably shallower than the location of the AAIW core. The location of convection in the western part of the South Pacific Ocean basin does not exclude the process of convection in this region from contributing to the formation of AAIW in the eastern part of the basin. It homogenizes the water column to depth of about 350 m and contributes to the formation of upper thermocline water (i.e. the mode waters that are transported eastward in the Antarctic Circumpolar Current). Along this pathway, vertical mixing may transport this water down into the AAIW level. In the CART run, large but extraneous diapycnal fluxes associated with horizontal mixing across too abruptly-sloped isopycnal surfaces in the Southern Ocean provide an additional pathway to the deeper density surfaces. These mechanisms have been elucidated in our passive tracer experiments. A comparison of both runs with observations from the South Pacific region indicate that while mid-latitude convection is exaggerated in the CART run it may be underestimated in the GM run. There are no direct observations of convective events in the Southern Ocean yet, but well-mixed layers to depths of 400-600 m are found in winter immediately north of the circumpolar Subantarctic Front (McCartney, 1977).

Rather than mid-latitude convection, subduction and along-isopycnal mixing along the sub-Antarctic front appear to be the dominating AAIW formation processes in coarse-resolution ocean models employing GM schemes and restoring boundary conditions at the ocean surface. These results are consistent with results from an isopycnal model (Marsh et al. 2000), which includes layer thickness diffusion analogous to the GM parameterization. Both studies indicate a circumpolar, subductive mechanism for AAIW formation, which locally may be enhanced by mid-latitude convection. Marsh (1999) also demonstrated that the ventilation of a subtropical gyre (in that study, it is the North Atlantic Ocean gyre) is sensitive to the choice of the thickness diffusion. By varying the interface diffusion velocity (equivalent to the GM thickness diffusion coefficient) Marsh (1999) found that with larger diffusion velocities, the gyre is progressively less well ventilated through shallower winter mixing. Since the same physics applies in both the isopycnal and z-coordinate model, excessive thickness diffusion may cause insufficient AAIW subduction. However, such sensitivity was not further investigated in the present study.

Our results contrast with those obtained in a study by England et al. (1993), where a direct source for AAIW in the eastern South Pacific Ocean was present and convection penetrated to about 800-1000 m depth. England et al. (1993) restored their model to zonally-averaged surface temperature and salinity values. Such salinity values are, however, biased toward the high surface salinity observed in the western Pacific Ocean. In turn, this would tend to weaken positive buoyancy fluxes or even lead to negative buoyancy fluxes in the model at mid-latitudes in the eastern South Pacific Ocean. This may explain deeper mid-latitude convection and penetration of convection down to the AAIW level in the England et al. (1993) model.

The inclusion of the GM scheme into OGCMs certainly improved the vertical structure of the simulated temperature and salinity distribution within the thermocline and reduced excessive convection in the high latitudes of the Southern Ocean. However, there is some indication that GM underestimates the depth of the mixed layer in the mid-latitudes where mode and intermediate water is formed. This is not only the case in the present study but was also found in previous model applications (Danabasoglu et al. 1994; England 1995; Duffy et al. 1997). Recent observations of the oxygen concentration in the Southern Ocean indicate mixed layer depths in excess of 300 m during winter (Hanawa and Talley 2001). This is significantly deeper than those depths simulated in the GM run. The observed oxygen data also allowed the origin of AAIW to be traced back to the eastern South Pacific surface waters (Talley 1996); a similar conclusion to that made McCartney (1977, 1982). This apparent inconsistency between model results and observations in regard to the presentation of the mixed layer requires further work in the future.

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Figure Captions

Figure 1: Globally-averaged profiles of (a) potential temperature (^oC), (b) salinity (psu), and (c) potential density (kg m^{-3}). Solid lines show the Levitus (1982) distributions, dashed lines show results from the GM run and dotted lines show results from the CART run

Figure 2: Vertical distribution of salinity (psu) as a Global Ocean average from the (a) CART and (b) GM runs, and the (c) Levitus (1992) climatology. Lowest salinity is shaded, contour interval is 0.1 psu.

Figure 3: Maximum annual convection depths (m) in the (a) CART and (b) GM runs. Convection depths are shown with contours of 50, 100, 200, 300, 500, 750, 1000, and 1250 m.

Figure 4: Annually-averaged surface buoyancy fluxes (kg m^{-2} $year^{-1}$) in the GM run: (a) total buoyancy flux as the sum of fluxes from heat (b) and fresh water (c) exchanges. Positive fluxes are shaded, contours are in intervals of 50 kg m^{-2} $year^{-1}$.

Figure 5: Salt transport (kg s^{-1}) into the intermediate water mass density range of the South Pacific Ocean control volume. Transport processes are: (1) total advection which includes the bolus transport; (2) bolus transport or isopycnal

advection (GM); (3) total diffusive mixing which includes vertical diffusive mixing; (4) vertical diffusive mixing (which includes the vertical component of isopycnal mixing in the GM run); and (5) convective mixing. The model is in a steady state, the total change is zero in both runs; i.e. the sum of (1), (3) and (5) is zero. The difference between (1) and (2) yields the non-isopycnal advection, which in the CART run is equal to (1) since only GM is characterized by an isopycnal component.

Figure 6: Equilibrium fraction of (top) SA and (bottom) NA tracer in a section across the Pacific at 30^o S from both the (a) CART and (b) GM runs. σ_t (kg m^{-3}) surfaces are overlaid.

Figure 7: Zonally averaged fractions of (a) SA and (b) NA tracer in the Pacific Ocean and for both the CART (left) and GM (right) experiment. Distributions are shown after 5, 25, 50 and 100 years. The SA tracer primarily ventilates the intermediate and deep ocean, while the NA tracer ventilates the upper thermocline.

Figure 8: (a) SA and (b) (NA) averaged tracer concentration (%) in the intermediate water mass range for both the CART (left) and GM (right) experiment. Distributions are contoured in intervals of 10 (%); the 5 (%) unit contour line is also shown.

TABLE 1. South Pacific Ocean volume transports in Sv [1 Sv = 10^6 m^3s^{-1}]. Transports were computed for a meridional section south of Australia, a meridional section across Drake Passage, and for a section across the South Pacific at 30^o S.

Density Range	150^o E		65^o W		30^o S	
	GM	CART	GM	CART	GM	CART
Surface Water σ_t<26.0	0	0	0	0	4.0	2.8
Upper thermocline Water 26.0<σ_t<26.8	23.2	34.0	9.1	12.4	6.4	10.4
Intermediate Water 26.8<σ_t<27.5	61.5	102.6	61.5	115.0	4.0	-0.1
Deep and Bottom Water 27.5<σ_t	53.3	23.1	49.5	13.3	3.7	6.4
Total	138.0	159.7	120.1	140.7	18.1	19.5